

THE COLD LOW OVER SOUTHEASTERN UNITED STATES, JULY 1-6, 1958

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1. INTRODUCTION

During the summer months anticyclonic circulation persists at the surface over southeastern United States. The airmass is usually either convectively unstable or in the process of becoming so, due to strong day-to-day insolation and the availability of moisture from the Gulf of Mexico and the Atlantic Ocean.

Because frontal systems and surface cyclonic circulation are lacking, showers and thunderstorms that occur in this airmass have been attributed to the release of conditional instability by surface heating regardless of the extent of the activity. Studies by Byers and Rodebush [1], and others, indicate that thermal instability, though a necessary condition, is not in itself sufficient cause for the development of widespread cloudiness and thunderstorm activity. Baum's study [2] indicated that the occurrence of thunderstorms in Florida cannot be determined purely from the standpoint of thermodynamics, even when the moisture parameter is taken into consideration. Widespread cloudiness and shower activity are always associated with relatively large-scale zones of convergence that are dynamically induced.

In this study an attempt will be made to explain on the basis of dynamically induced convergence the cloud and precipitation pattern that developed with the upper cold Low circulation over southeastern United States, July 1-6, 1958.

2. ANTECEDENT CONDITIONS

During the last week of June a cold front moved across the eastern United States. The portion of this cold front that had extended across northern Florida dissipated and by the last day of the month a high pressure system dominated the weather over southeastern United States. The air behind the cold front was a comparatively cool and dry Canadian type.

Aloft, at 500 mb. and 250 mb., a large-amplitude trough associated with the surface frontal system extended from a Low centered near Sable Island south-southwestward across the extreme western Atlantic Ocean, across the northern border of Florida and into the Gulf of Mexico. With the zone of strong westerlies extending across extreme southern Canada, the portion of the trough north of 36° N. continued moving eastward. The southern portion, with generally light circulation, tended to remain stationary.

This shearing of the trough was evident at 0000 GMT, June 30. Warm air advection and generally rising heights were indicated over the Northeast and the northernmost Mid-Atlantic States.

3. DEVELOPMENT

As the northern portion of the trough sheared eastward, warm air advection aloft across New England and the northernmost Mid-Atlantic States brought height rises over these areas. This was evident at 500 mb. (fig. 1 A and B) and to a greater degree at 250 mb. (fig. 2 A and B).

As a result of building of the 250-mb. heights to the north of the stationary portion of the trough with little change to the south, there was an increase of cyclonic vorticity south of the height rises. Winds over Virginia and North Carolina veered to the east and a closed low circulation was established over extreme southeastern United States by 1200 GMT, July 1.

Through the period of study (fig. 2 A-F), the low center drifted first to the southwest across extreme northwestern Florida and out over the northern Gulf of Mexico. It then followed a recurving path northward over Alabama, then east-northeastward as it became imbedded in the westerly flow across central United States. The Wilson [3] grid method was tried in forecasting this movement with good to fair results. Surface circulation during the entire period remained weak and anticyclonic.

4. STABILITY

Stability index charts, computed by the Showalter [4] method for obtaining a measure of conditional instability (fig. 3 A-F), indicated that in general the air over southeastern United States was convectively stable during the first day of the period. This is not surprising in view of the Canadian origin of this air. The air over southern Florida, a mixture of Canadian and tropical maritime airs, was conditionally unstable.

The changes in stability index, progressing through the period, are attributed, in part, to heating and the addition of moisture in the low levels of the relatively cool air, and to cold air advection associated with the moving cold Low in the high levels. Comparison of the soundings on the vertical cross sections (figs. 4 and 5) shows that this differential advection, heating below and cooling aloft,

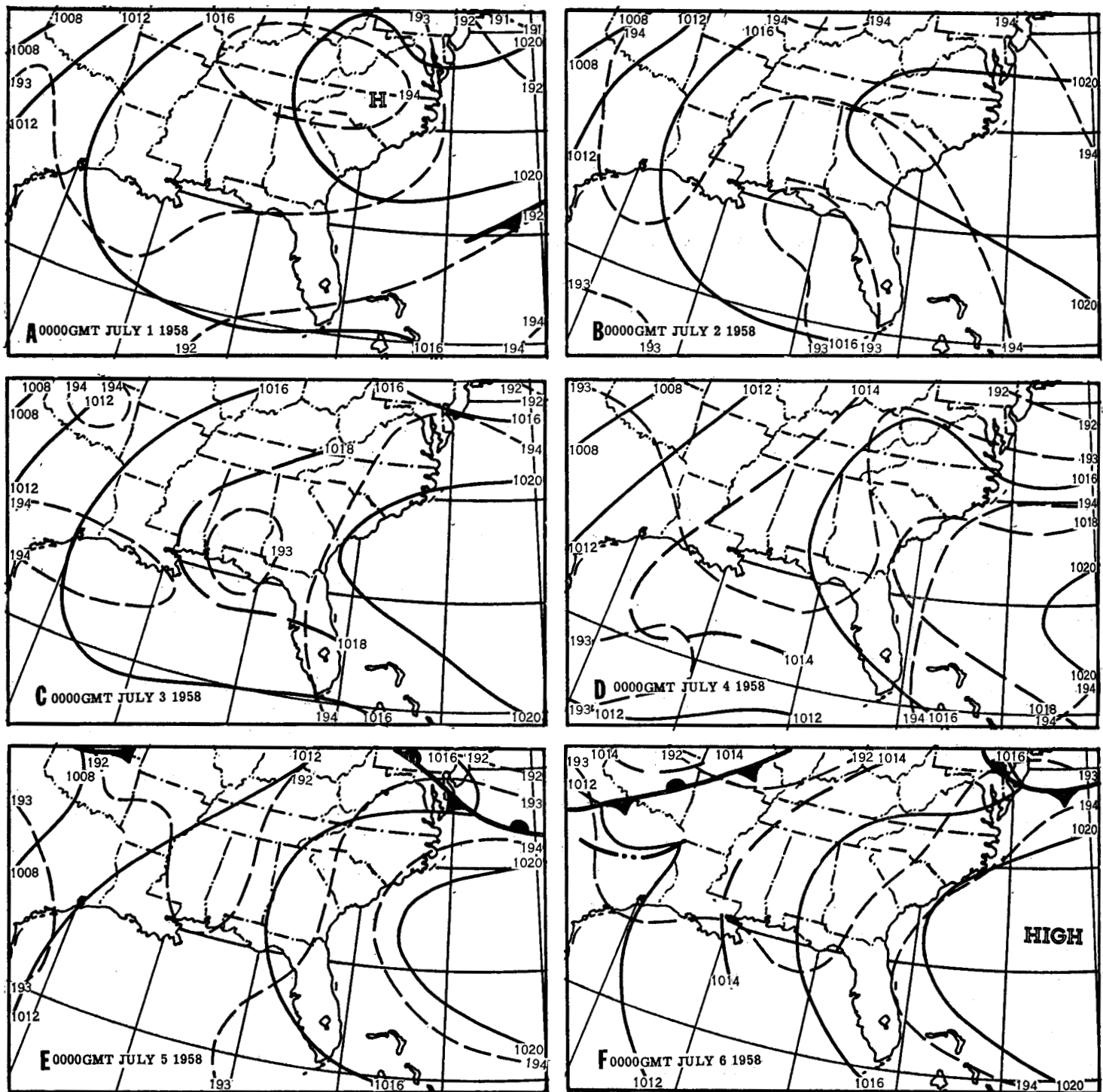


FIGURE 1.—Sea level isobars (solid lines) and 500-mb. contours (dashed lines) for 0000 GMT, July 1 to July 6, 1958. 500-mb. contours are in hundreds of geopotential feet.

resulted in packing of the isotherms and indicated decreasing stability. In the vertical cross sections, Bellamy D values [5] are used to illustrate the vertical structure of the atmosphere. The rate of change of D values also indicates stability condition; i. e., cooling is depicted by increasing negative departures or decreasing positive departures.

The stability index at Montgomery, Ala., decreased from +3, July 2 to -2, July 4, as the cold Low moved from just south of Panama City, Fla., to near the Mobile, Ala., area. Jackson, Miss., during the same period, showed a stability index decrease from +4 to +1.

The change in stability index at Montgomery resulted from an increase in temperature and moisture at the

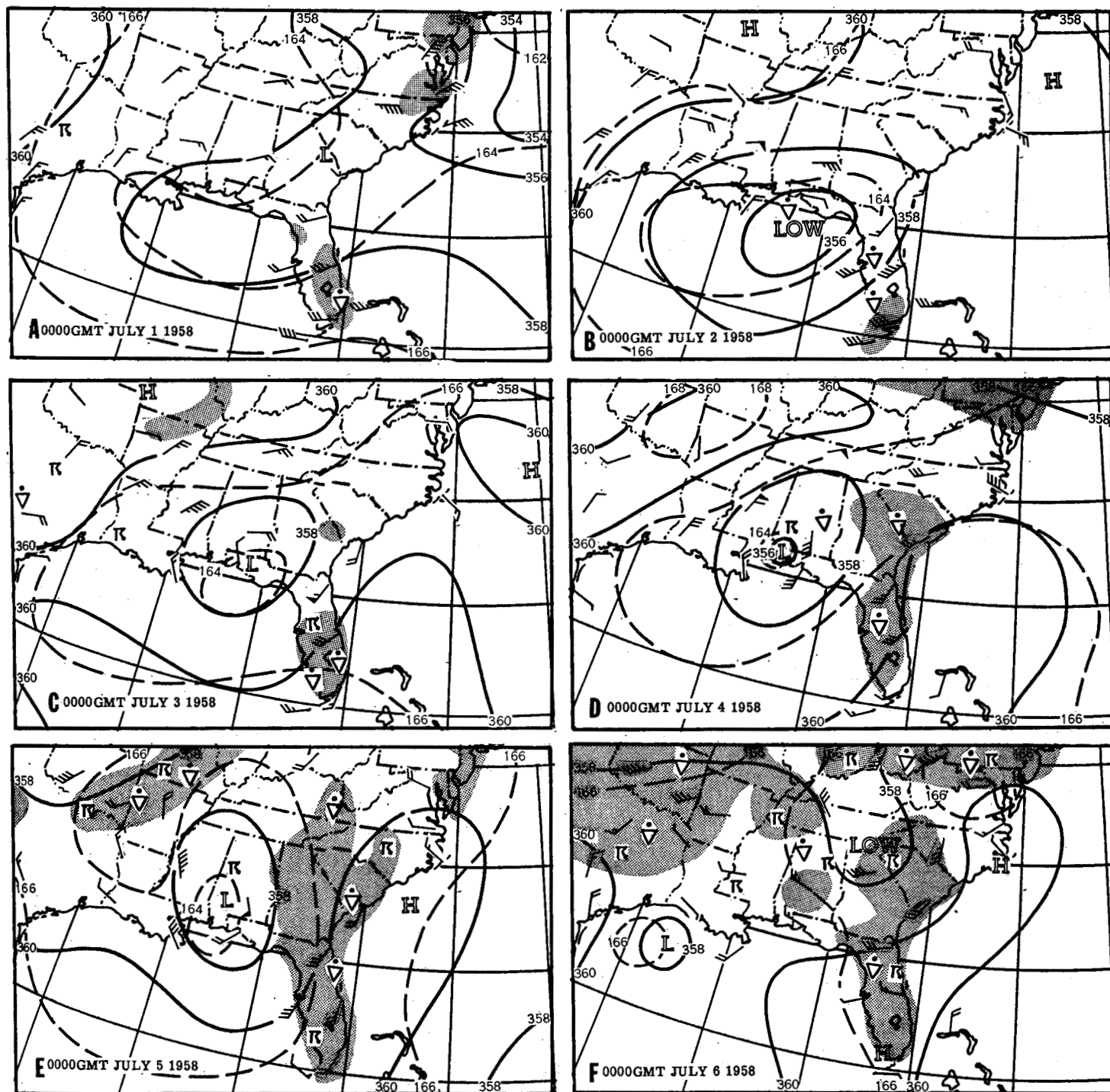


FIGURE 2.—250-mb. maps for 0000 GMT, July 1 to July 6, 1958 with 500–250-mb. thickness lines (dashed). Current cloudiness is shown by shading, and areas of precipitation by symbols.

850-mb. level and cooling at the 500-mb. level. At Jackson, the change was due only to an increase in temperature and moisture at the 850-mb. level since the 500-mb. temperature was unchanged.

Farther to the east of the upper low center, Athens, Ga., showed a decrease in stability index from +4 to -1, attributable to increased temperature and moisture at

the 850-mb. level with no change in temperature at the 500-mb. level. Lake Charles, La., to the west of the Low, showed a slight increase in stability.

It is noteworthy that to the east of the Low (figs. 4 and 5) marked moisture increases were indicated up through the 500-mb. level, whereas to the west the increases were slight and occurred only in the lowest levels.

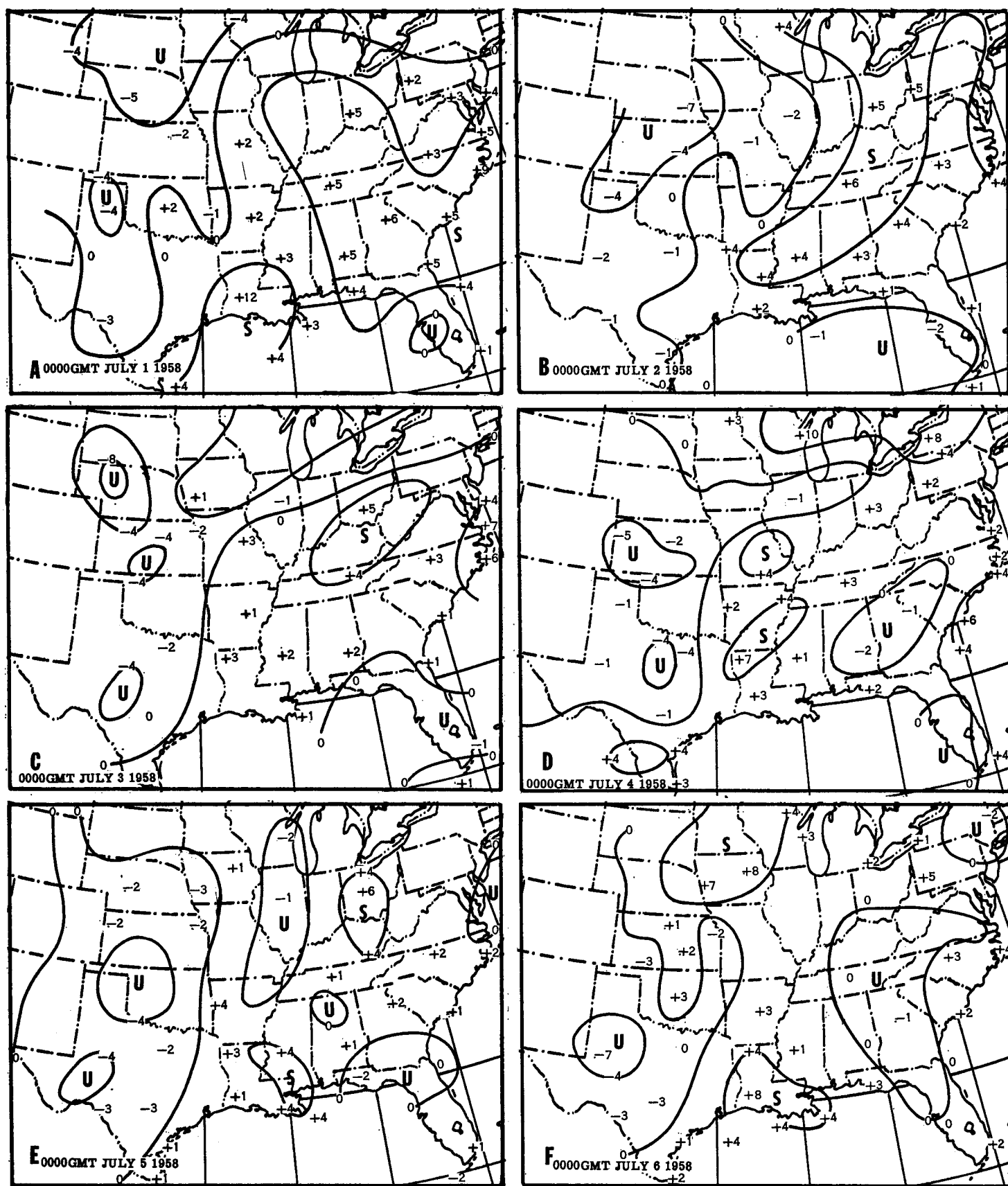


FIGURE 3.—Showalter stability index for 0000 GMT, July 1 to July 6, 1958.

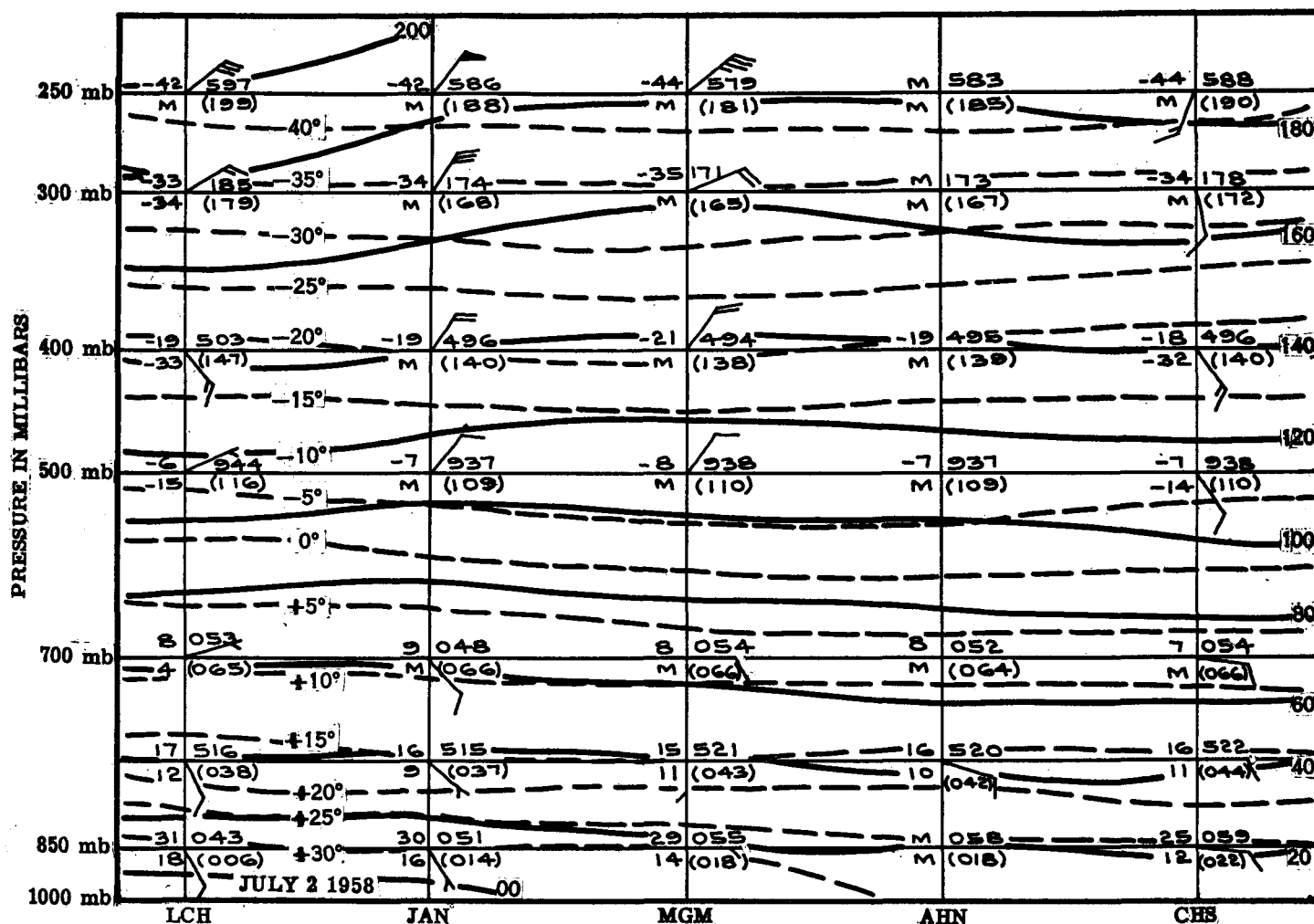


FIGURE 4.—Vertical cross section at 0000 GMT, July 2, 1958. Figures in parentheses are departure from standard atmosphere (D values). Isolines of D values (solid) are in hundreds of feet. Isotherms (dashed lines) are at intervals of 5° C. LCH=Lake Charles, La., JAN=Jackson, Miss., MGM=Montgomery, Ala., AHN=Athens, Ga., CHS=Charleston, S. C.

The generally weak circulation at the 500-mb. and lower levels and the fact that the air was initially fairly homogeneous rule out the possibility of horizontal advection as the cause of this increase in moisture. Since the only likely source of moisture was the surface, the observed changes indicate the extent of vertical motion.

Vertical motion has the effect, other than of transporting moisture aloft, of stretching the air column in the vertical and decreasing absolute stability. Vertical motion cannot be determined directly from observations but can be related to vorticity and divergence. The Joint Numerical Weather Prediction Unit was preparing vertical motion charts during this period while experimenting with a two-level baroclinic model. These charts (not shown) indicated a weak positive vertical velocity over all of southeastern United States and, of course, did not explain the variation in distribution of moisture aloft. This could be because of the general weakness of the circulation up through the 500-mb. level and the result of smoothing out of small-scale features which can be

locally important. In the x, y, p coordinate system the vertical p -velocity, $\omega = dp/dt$, is related to wind divergence, $\text{div}_p \mathbf{V}$, measured on an isobaric surface by the equation of continuity written in the form (c. f., Panofsky [6]):

$$\frac{\partial \omega}{\partial p} = -\text{div}_p \mathbf{V}$$

Although it is possible to evaluate divergence from reported winds, this is difficult because the greatest contribution to divergence is made by wind deviations from the geostrophic wind. For the purpose of this discussion, only the sign of the vertical motion is required to explain its contribution to change in stability and the weather.

We can approximate the vertical distribution of divergence from the vorticity theorem and from this distribution and the equation of continuity we can determine the sign of the vertical motion. (Note that although ω and w , the real vertical motion, have opposite signs, $\partial \omega / \partial p$ and $\partial w / \partial z$ have the same sign.) The isobaric horizontal

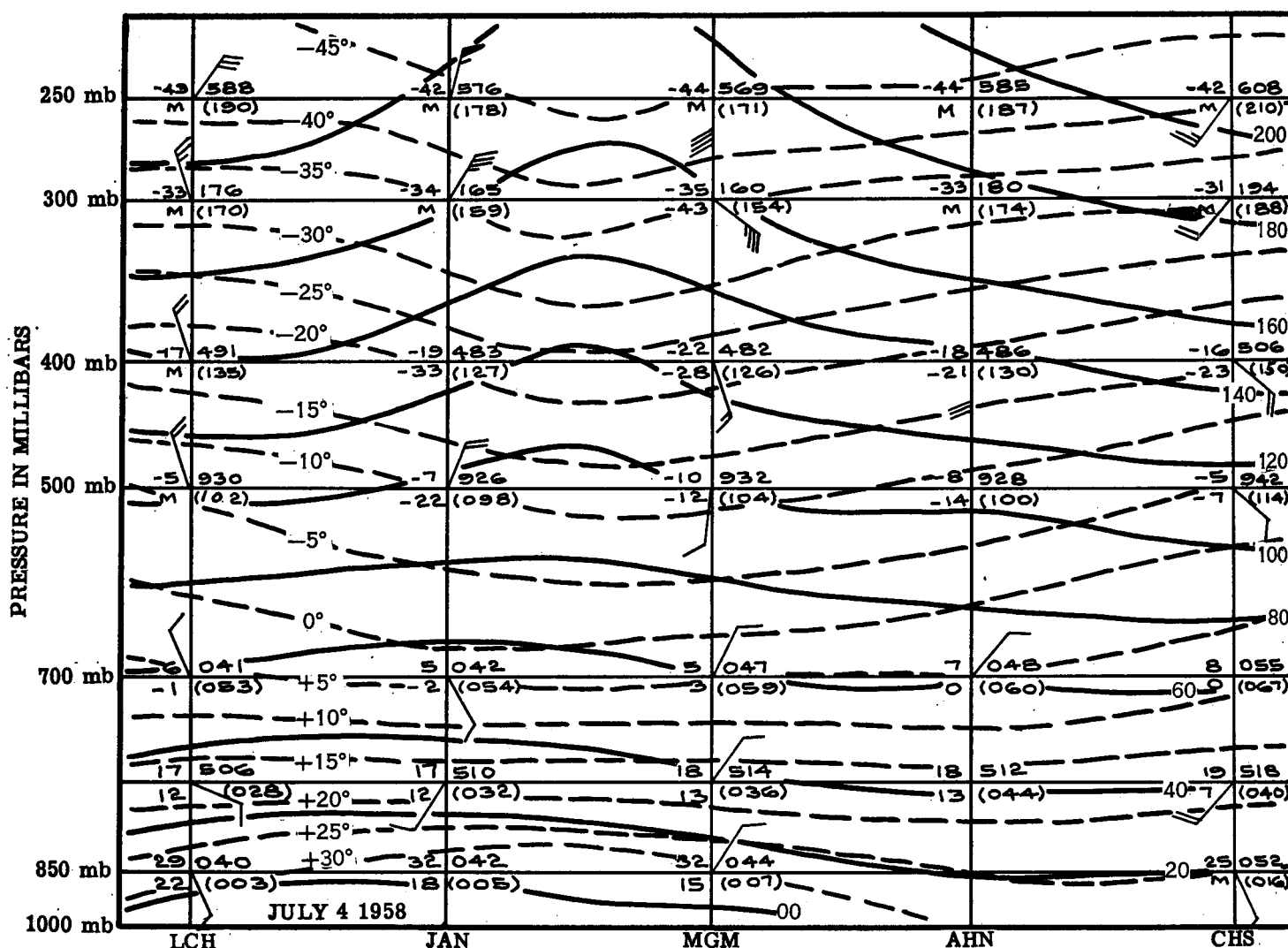


FIGURE 5.—Vertical cross section at 0000 GMT, July 4, 1958.

divergence $\text{div}_p \mathbf{V}$ is given by the vorticity equation written in the form:

$$\text{div}_p \mathbf{V} = -\frac{1}{\zeta + f} \left(\frac{d\zeta}{dt} + \beta v \right)$$

where ζ is the relative vorticity, f the Coriolis parameter, t time, β the rate of change of f along the y axis, and v the component of the wind along the y axis (positive northward).

If the relative vorticity is negative at ridges and positive at troughs, and the air moves faster than the system (Panofsky [6]), $d\zeta/dt$ will be negative east of troughs and positive east of ridges. The βv term is positive east of troughs and negative east of ridges but is relatively less important than $d\zeta/dt$ at high speeds. At high levels, therefore, we can expect divergence east of troughs and conversely convergence east of ridges. At low levels, with slower winds, the βv term becomes more important; if the winds are slower than the system, $d\zeta/dt$ has the same

sign and therefore also contributes to convergence ahead of troughs and divergence east of ridges. Hence, from the equation of continuity, we can expect ascending motion east of the upper trough and descending motion west of the trough.

Horizontal divergence charts for July 4 (figs. 6 and 7), prepared as described by Saucier [7], give the distribution of the horizontal divergence at 2,000 ft. and the 250-mb. level, and agree favorably with what has been discussed. The resulting vertical motion explains the occurrence of the unstable areas (fig. 3 A-F) east of the low center and the tendency for stabilization to the west.

The following paragraph discusses the association of weather with vertical motion and changes in stability.

5. PRECIPITATION

During the first day of the period, precipitation, in the form of showers, and cloudiness were confined to the unstable area over southern Florida (fig. 2A). By July 4

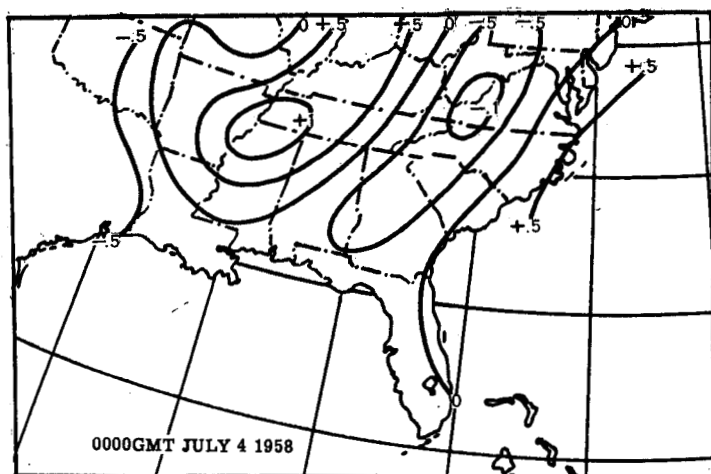


FIGURE 6.—Computed horizontal divergence (10^{-5} sec^{-1}) for 2000-ft. level at 0000 GMT, July 4, 1958. Positive areas indicate horizontal divergence and negative areas horizontal convergence.

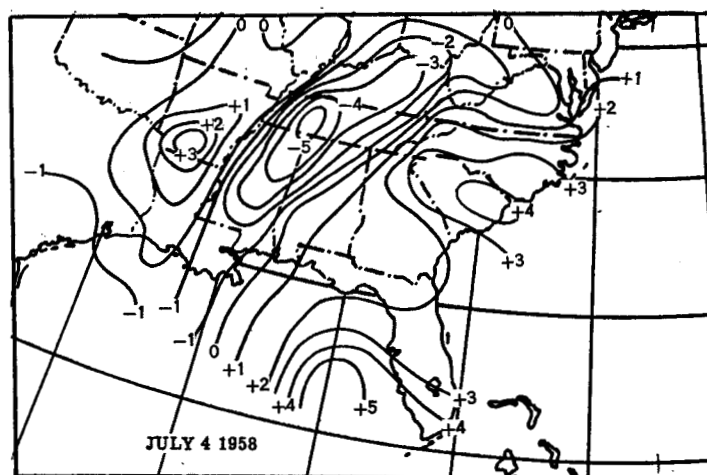


FIGURE 7.—Computed horizontal divergence (10^{-5} sec^{-1}) for 250-mb. level at 0000 GMT, July 4, 1958.

(fig. 2D), cloudiness and shower activity had spread over all of northern Florida, most of Georgia, and South Carolina, but west of the Low clear skies continued and no precipitation was reported. It is evident that the cloud and precipitation pattern obtaining through the period was closely related to the vertical motion and the development of instability to the east of the low center and stability to the west. Showers and thunderstorms reported in areas other than those enclosed in figure 2 appeared to be widely scattered.

6. CONCLUSION

The review of this storm brought forth an appreciation of the effects of dynamically induced divergence and vertical motion on stability and the weather. These effects and the weather could only be explained when the comparatively stronger high-level cyclonic circulation was considered, since at the surface the circulation remained weakly anticyclonic during the entire period. This dynamic lifting served the twofold purpose of increasing and releasing convective instability.

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